THE LOWER IONOSPHERE OF MARS

by

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APRIL 1968

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ABSTRACT

A study is made of the ionization of the Martian atmosphere below 100 km in the context of recent data on the atmosphere and ionosphere of the planet. Suggested sources of ionization include: solar ultraviolet and x-radiation, cosmic rays, meteors, radioactive material within the atmosphere and radioactivity of the surface, the solar wind, and protons originating from solar flares. Loss processes, including charge rearrangement, dissociative recombination of ions and electrons, and negative ion formation are detailed. Possible importance of the negative ion CO_4^- is discussed. Its distribution will depend on the O_2 content of the atmosphere as well as the distribution of O and CO. CO^- and CO_3^- may also be present in the ionosphere. It is shown that combination of the solar photoionization rate with that due to cosmic radiation will lead to an ion density minimum. The magnitude of the ion density and location of the minimum will depend on the atmospheric model chosen, solar activity, and solar zenith angle as well as such factors as the presence of particulate matter and meteor ionization.

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INTRODUCTION

Recent experimental results have yielded much definite information on the properties of the Martian atmosphere. In particular spectroscopic studies have given us a more complete picture of the chemical composition, temperature, and pressure distribution with altitude, Gray (1966), Belton and Hunten (1966). Similar information has been derived from the Mariner IV occultation experiment conducted during the fly-by of Mars in July 1965, Kliore et al (1965). Mariner IV also yielded important data concerning the ionosphere of Mars, Fjeldbo et al (1966a,b), Kliore et al (1965). The electron concentration distribution as derived from the Mariner IV 2 frequency occultation experiment is shown in Figure 1, Fjeldbo et al (1966a). An electron density scale height of 24 ± 3 km was found together with a maximum electron density of $9.5 \pm 1.0 \times 10^4$ el/cm^3 at an altitude of $120 + 5$ km. The plasma height remained constant to an altitude of 250 km. There is a minor ionospheric layer at 100 km with a maximum electron density of 2.5×10^4 cm^{-3}. An upper limit of 5×10^3 cm^{-3} was found for the nighttime ionosphere, Kliore et al (1965), Fjeldbo et al (1966a).

Several atmospheric models have been proposed to explain the occultation data. These can be categorized according to their description of the ionosphere as analagous to the Earth's ionospheric E, F-1, and F-2 regions as regards to the formation. The designations are interpreted to mean that in the E-region case the ionospheric peak is formed by a balance the ion pair production rate due to long wavelength solar x-rays with recombination of CO_2^+ ions, Chamberlain

and McElroy (1966), McElroy (1967). The primary ion is considered to be O^+ at the observed electron density maximum of 120 ± 5 km for the F-1 region, Donahue (1966). Diffusion of ions and electrons to form the peak is the basis of the F-2 region, Johnson (1965).

The present paper is a study of the lower ionosphere of Mars, i.e., that portion which lies between the surface of the planet and an altitude of approximately 100 km. The description of such an ionosphere will depend on the atmospheric model assumed. For comparison, two atmospheric models are employed, namely, the Johnson "F-2" and the Chamberlain and McElroy "E". It will be shown that an F-1 type atmosphere gives rise to a lower ionosphere between these two extremes.

SOURCES OF IONIZATION

The following sources of ionization must be considered for the atmosphere of Mars:

1. Solar ultraviolet and x-radiation

2. Cosmic rays

3. Meteors

4. Natural radioactivity of the planetary surface and radioactive material within the atmosphere.

5. The solar wind

6. Protons originating from solar flares

Since the planetary magnetic field is at most 3×10^{-4} times that of the Earth, Smith et al (1965), no radiation belts and auroral particle precipitation

zones are to be expected. Indeed Van Allen et al (1965) and O'Gallagher and Simpson (1965) found no evidence of trapped radiation.

In considering the effect of solar radiation it is essential to know the intensity as a function of wavelength, as well as the cross sections for absorption and photoionization of the gases constituting the Martian atmosphere. The intensity of the solar radiation at 1AU has been summarized by Hintereger, et al (1965). The intensity at distances other than 1AU can easily be calculated knowing the properties of the orbit of Mars. The distance of the planet from the sun varies between 1.36 and 1.67 AU over a two year period. The mean distance is 1.52 AU. This leads to a dilution factor of between 0.524 and 0.36, Schultz and Holland (1962). At the time of Mariner IV the planet was in a position such that the dilution factor was 0.39.

As stated previously the possible constitutents of the Martian atmosphere are CO_2, N_2, Ar,O,CO and possibly O_2 and oxides of nitrogen such as NO. The wavelength equivalent of the ionization potential of these constitutents is given in Table I.

Table I

Ionization Thresholds of Martian Atmospheric Constituents

Ar	787A
N_2	796A
CO	885A
CO_2	899A
O	910A
O_2	1026A
NO	1340A

can be overcome if an additional ionization source exists. One source not previously considered is that of meteor ionization. In the Earth's atmosphere most meteors deposit their energy between 90 and 120 km with some coming as low as 70 km. For an incoming meteor of mass m, density ρ_m, and velocity V, the number of ion-pairs produced per unit path length is according to McKinley (1961)

$$q = \tau_q \, B \, \frac{m^{2/3}}{\rho_m} \, \rho_a \, V^4$$

where τ_q is the ionization efficiency factor B is a constant which is a function of the shape and ablation properties of the meteor and ρ_m is the density of the atmosphere. Assuming that the atmospheric density distribution is described by a constant scale height H the maximum ionization will occur where

$$\rho_a \, H = C \, m^{1/3} \, \rho_m^{2/3} \, V^{-2} \, \cos \chi$$

Here the path length (ds) in terms of the verical height, dz, is

$$ds = dz/\cos \chi$$

and C is another constant which is a function of B. Table II gives the altitude of maximum ionization for the different Martian atmosphere models in comparison with deposition altitudes for equivalent meteors impinging on the Earth's atmosphere.
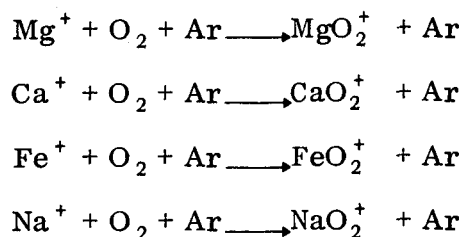
Some of the most intense line and continuum emissions of the solar ultra-violet spectrum occur at wavelengths longer than 900A. This radiation includes the important Lyman series Lyman alpha (1215.7), Lyman beta (1025.7), etc. and the Lyman continuum (910A) which will be almost ineffective in producing ionization unless NO, O_2, and O are present in the atmosphere. For an atmosphere composed principally of CO_2, HeI (584.3A) and HeII (303.8A) as well as continuum radiation below 200A will play an important role.

Figure 2 illustrates the ion-pair production functions due to solar radiation for the McElroy E and F-1 models and the Johnson F-2 model for a solar zenith angle of 67°. The latter was calculated by the author; the E and F-1 models are reproduced from McElroy (1967). The peak production rates correspond in all cases to unit optical depth for wavelengths in the vicinity of 304Å. The second-ary peaks at lower altitudes correspond in the E and F-1 region cases long wavelength x-rays. For the F-1 model the peak of production should corre-spond to the peak of the electron density distribution provided the loss processes are not altitude dependent. McElroy (1967) has demonstrated that in the case of the F-1 layer, the maximum of ionization occurs below the production peak. The same thing occurs in the case of the E region, the 120 km peak being caused by solar radiation at x-ray wavelengths, where the flux is extremely variable. The F-2 model has the peak of production near 100 km. Such a model requires a loss rate in excess of 10^{-6} cm^3 sec^{-1} to explain the observed low altitude variation.

The problem of the need for an additional ionization source at low altitudes which is most acute for the E region model, but exists for the F-1 model also,

dust should be at least as probable on Mars as on the Earth. The actual presence of meteoric layers (already established for the Earth's atmosphere) will depend on the loss process acting in the Martian atmosphere.

In addition to direct ionization by meteor ablation, photoionization of accumulated meteoric atoms can occur. These atoms are primarily Fe, Mg, Si, Ca, and Na. Their presence in ionized form has been detected in the Earth's atmosphere at the turbopause by rocket-borne ion mass spectrometers both during ordinary and meteor shower conditions. Narcisi et al (1967 a,b). However, estimates of the ionization rate coefficients were given many years previously by Nicolet (1955). These rates $I(Na) = 10^{-5}$ sec^{-1}, $I(Ca) = 10^{-6}$ sec^{-1}, $I(Mg) = 10^{-7}$ sec^{-1} can be used to estimate the number of atoms necessary in order to yield an ion density of 10^3 cm^{-3}, assuming a radiative recombination coefficient of 10^{-12} cm^3 sec^{-1}. The density of atoms is less than 10 cm^{-3}. Fehsenfeld et al (1968) have given rate coefficients for the following processes

$$Mg^+ + O_2 + Ar \longrightarrow MgO_2^+ + Ar$$

$$Ca^+ + O_2 + Ar \longrightarrow CaO_2^+ + Ar$$

$$Fe^+ + O_2 + Ar \longrightarrow FeO_2^+ + Ar$$

$$Na^+ + O_2 + Ar \longrightarrow NaO_2^+ + Ar$$

Similar reactions may occur in the case where CO_2 is substituted for O_2. Such reactions or an atmosphere containing sufficient O_2 could lead to molecular ions. These ions will dissociatively recombine with electrons with a rate coefficient in the 10^{-6} to 10^{-7} cm^3 sec^{-1} range. Under such circumstances metalic atom densities in the 10^6 to 10^7 cm^{-3} range are necessary in order to produce 10^3 electrons cm^{-3}.

Table II

Deposition Height of Meteors

Earth Altitude	McElroy E	McElroy F-1	Johnson F-2
70	74	61	44
80	91	74	55
90	111	89	65
100	129	102	73
110	145	113	80

Various estimates of the contribution of meteors to the ion-pair production function have been made for the earth and range in value between 10^{-5} and 10^{-1} electrons cm^{-3} sec^{-1} e.g. Lovell (1957). Fiocco (1967) has reexamined the problem considering recent information on the interaction of hypersonicly moving bodies with the Earth's atmosphere. The ionization rate depends on the area/mass ratio of the particle as well as its velocity, the efficiency of ionization and flux of incoming particles. For velocities between 39.2 and 44.7 km/sec and area/mass ratios between 10 and 10^3 m^2/kgm, the production function varies between 10^{15} and 10^{18} ion pairs/kgm m with values between 10^{17} and 10^{18}/kgm m being most typical.

An average value for the incoming flux at the Earth is 10^{-13} kgm/m^2 sec, which could lead to an ion-pair production function of 10^{-5}/cm^3. Dust measurements by Alexander et al (1965) conducted on Mariner IV indicated a factor of five increase in the flux at 1.4 AU as compared with that at 1 AU, although there was a slight decrease in the immediate vicinity of Mars. Thus ionization by

An ionization chamber was included as part of the instrumentation of Mariner IV. The instrument ceased operation on 16 February 1965. One can compare this data with that obtained by the Mariner II probe to Venus in 1962. The data, taking into account the masking effect of the planet, gives 330 ion pairs/cm^3 sec atm for 1962, Neher and Anderson (1964), and 500 ion pairs/cm^3 sec atm for 1965, Neher as reported by Nicks (1967). Assuming the ionization efficiency of CO_2 to be the same as air, the ionization rate for the time of Mariner IV can be written as

$$q_{CR}(z) = 1.85 \times 10^{-17} n(M) \text{ ion pairs/cm}^3 \text{ sec}$$

where n(M) is the local number density at any altitude, Z. Figure 2 illustrates the variation of q_{cr} as a function of altitude for the different atmospheric models.

The undisturbed solar wind is composed of protons of velocity averaging between 300 and 400 km/sec (500 to 900 ev) with densities up to 20/cc (Parker, 1964). Because of the absence of a magnetic dipole of strength greater than 10^{-3} of the Earth; the solar wind will penetrate the Martian atmosphere. Indeed Sagan and Veverka (1967) explain the subsidiary ionospheric peak as being caused by the solar wind impinging on an F-2 type atmosphere. Recent information from the Mariner V probe to Venus give strong evidence that the solar wind does not penetrate the atmosphere. These results indicate an ionospheric conductivity-solar wind magnetic field interaction which causes the solar wind to slip around the planet. The situation should be similar for Mars, Mariner Stanford Group, (1967), Lazarus et al (1967).

Table III illustrates the altitude to which more energetic protons will penetrate for different Martian atmospheric models. Solar flare protons will dissipate most of their energy near the surface of the planet.

Table III

Depth of Penetration of Protons into the Martian Atmosphere

Model	10 Kev	100 Kev	1 Mev	10 Mev	50 Mev
Johnson F-2	80 km	72 km	57 km	35 km	11 km
McElroy F-1	103 km	88 km	70 km	35 km	10 km
McElroy E	140 km	125 km	95 km	50 km	15 km

As a result of the natural abundance of the elements, it would be expected that the ion-pair production due to the decay of elements within the Martian surface should be comparable to that of the Earth. It has been noted previously that the main production of cosmic ray secondaries is below the Martian surface. This may increase the radioactivity of the surface. However, for our purposes we shall assume that the ion-pair production function is the same as that of the Earth. Bricard (1965) has given an excellent review of the role of radioactivity in atmospheric electricity and estimates that the contribution of the surface is $3.5 + 2.5 - 1.5$ ion-pairs/cm^3 sec/atmosphere.

According to Bricard (1965) (Table I after Israel (1961)), 0.1% of the soil radiation remains at 1 km above the surface of the Earth. This corresponds to a range in excess of 100 gm/cm^2, surface radiation will suffer little attenuation in its passage through the atmosphere of Mars. At the same time it represents only about 1% that which is produced by cosmic radiation. At this point it is not

possible to speculate on the radioactivity content of the Martian atmosphere. Radioactivity aerosols may well exist.

POSITIVE ION REACTIONS

The principal positive ion reactions which will occur in the Martian ionosphere are listed in Table IV together with their rate coefficients and sources. Also listed are the important positive ion- electron reactions which should be considered. For an atmosphere where O^+ is the predominant ion produced by photoionization CO_2^+ will result from the reaction of O^+ with CO_2. Rapid charge exchange will occur between N_2^+ and CO_2. Since the dissociative recombination coefficient of CO_2^+ as measured by Weller and Biondi (1967) is $3.8 \pm 0.5 \times 10^{-7}$ $cm^3 \ sec^{-1}$ at 300°K and of the same order at 210°K, O_2 concentrations in excess of $4 \times 10^8 \ cm^{-3}$ must be present for electron densities of $10^5 \ cm^{-3}$ or greater in order for O_2^+ to become the dominant ion when reactions between CO_2^+ and O_2 occur.

Table IV

Important Positive Ion Reactions in the Martian Ionosphere

Reaction		Rate $cm^3 \ sec^{-1}$		
$Ar^+ + CO_2 \rightarrow Ar + CO_2^+$	(1)	7.4×10^{-10}	$C^+ + CO_2 \rightarrow CO + CO^+$ (8)	1.8×10^{-9}
$Ar^+ + CO \rightarrow Ar + CO^+$	(1)	8.0×10^{-11}	$C^+ + O_2 \rightarrow O + CO^+$ (8)	1.1×10^{-9}
$Ar^+ + O_2 \rightarrow Ar + O_2^+$	(1)	1.1×10^{-10}		
$Ar^+ + N_2 \rightarrow Ar + N_2^+$	(1)	6×10^{-12}		
$CO^+ + CO_2 \rightarrow CO + CO_2^+$	(8)	1.1×10^{-9}		
$CO^+ + O_2 \rightarrow CO + O_2^+$	(8)	2×10^{-10}		

Table IV (Continued)

Important Positive Ion Reactions in the Martian Ionosphere

| <u>Reaction</u>		<u>Rate</u>
$N_2^+ + CO_2 \longrightarrow N_2 + CO_2^+$		8×10^{-10}
$N_2^+ + O_2 \longrightarrow N_2 + O_2^+$	(2)	1×10^{-10}
$N_2^+ + O \longrightarrow N + NO^+$	(3)	2.5×10^{-10}
$O_2^+ + N_2 \longrightarrow NO^+ + NO$	(2)	$> 10^{-15}$
$CO_2^+ + O_2 \longrightarrow CO_2 + O_2^+$		1.0×10^{-10}
$CO_2^+ + NO \longrightarrow CO_2 + NO^+$		—
$O^+ + CO_2 \longrightarrow CO + O_2^+$	(8)	1.2×10^{-9}
$O^+ + O_2 \longrightarrow O_2^+ + O$	(2)	2×10^{-11}
$O^+ + N_2 \longrightarrow N + NO^+$	(2)	4.6×10^{-12}
$CO_2^+ + e \longrightarrow CO + O$	(4)	4×10^{-7}
$O_2^+ + e \longrightarrow O + O$		$5.2 \times 10^{-6}/T_e^{0.5}$
$N_2^+ + e \longrightarrow N + N$	(5)	$2 \times 10^{-6}/T_e^{0.33}$
$NO^+ + e \longrightarrow N + O$	(6)	$1.5 \times 10^{-4}/T$
$CO^+ + e \longrightarrow C + O$	(7)	6.8×10^{-7}
$Ar^+ + e \longrightarrow Ar + hv$		1×10^{-12}
$O^+ + e \longrightarrow O + hv$		1×10^{-12}
$C^+ + e \longrightarrow C + hv$		1×10^{-12}
$Mg^+ + e \longrightarrow Mg + hv$		1×10^{-12}

1) Warneck (1967a); 2) Warneck (1967b); 3) Ferguson et al (1965); 4) Weller and Biondi (1967); 5) Mehr and Biondi (1968); 6) Weller and Biondi (1968); 7) Mentzoni and Donohoe (1968); 8) Norton et al (1966).

The measured rate of the CO_2^+ dissociative recombination coefficient, 4×10^{-7} cm^3 sec^{-1} does not support the assumption of 10^{-5} cm^3 sec^{-1} made by McElroy (1967) in order to obtain agreement between the E region model and the experimental data. The assumption of a positive exponent temperature dependence of α_{O_2} (10^{-8} cm^3 sec^{-1} at 330°K and 10^{-7} at 300°K) is also in disagreement with laboratory data.

In contrast the F-2 model of Johnson requires that the α_{O_2} be of the order of 10^{-5} cm^{-3} sec^{-1} where the E region would have it be 10^{-8} cm^{-3} sec^{-1}. In the present work it will be assumed that α_{O_2} follows the temperature dependence given in Table IV and that α_{CO_2} has a $T^{-0.5}$ or $T^{-1.2}$ dependence. This choice is consistent with laboratory measurements.
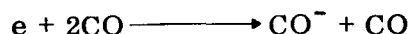
NEGATIVE ION REACTIONS

Laboratory experiments have not found direct attachment of electrons to CO_2 at thermal energies, Pack et al (1962) and Schlumbohm (1962). Attachment has been found for electron energies in excess of 2 ev, Schulz (1962). Ion mass spectrometer analysis has been used by Moruzzi and Phelps (1966) to establish the attachment process at these ionosphericly large electron energies. The reactions

$$e + CO_2 \longrightarrow CO + O^-$$

and

$$O^- + 2CO_2 \longrightarrow CO_3^- + CO_2$$
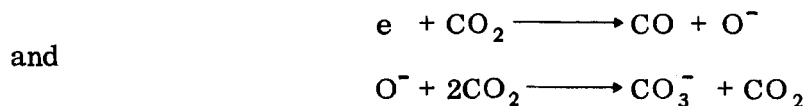
were found to occur. The ion CO_3^- is found to be dominant and would be expected in a pure CO_2 atmosphere with low energy electrons.

Several other species are possibilities for trace amounts of constituents common to the atmosphere of the Earth. One such example is the formation of CO^- which will occur by the three-body process involving thermal electrons

$$e + 2CO \longrightarrow CO^- + CO$$

at a rate of 2.7×10^{-31} cm^3 sec^{-1} Mentzoni and Donohoe (1968). The reactions resulting in the loss of this ion are unknown at present. However, this ion is a possibility if CO is present in the lower atmosphere of Mars.

An example of a negative ion distribution can be given for an atmosphere which contains O_2. The three-body attachment of thermal electrons to form O_2^- is well established, Chanin, Phelps, and Biondi (1962). This process
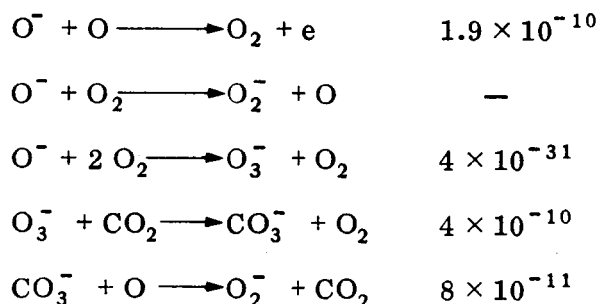
$$e + O_2 + M \longrightarrow O_2^- + M$$

is followed by

$$O_2^- + CO_2 + M \longrightarrow CO_4^- + M$$
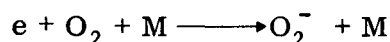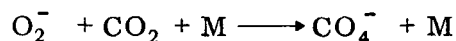
Pack and Phelps (1966a) have determined that when M is CO_2, the rate coefficient is 2×10^{-30} for the attachment process. Mourzi ahd Phelps (1966) have deduced a rate of $9 \times 10^{-30} cm^6/sec$ for CO_4^- formation where M is CO_2. If M is O_2, this rate coefficient is 2×10^{-29} cm^6/sec.

Other charge exchange reactions which can occur in O_2, CO_2 mixtures include

$$O^- + O \longrightarrow O_2 + e \qquad 1.9 \times 10^{-10}$$

$$O^- + O_2 \longrightarrow O_2^- + O \qquad -$$

$$O^- + 2O_2 \longrightarrow O_3^- + O_2 \qquad 4 \times 10^{-31}$$

$$O_3^- + CO_2 \longrightarrow CO_3^- + O_2 \qquad 4 \times 10^{-10}$$

$$CO_3^- + O \longrightarrow O_2^- + CO_2 \qquad 8 \times 10^{-11}$$

For an atmosphere having O as a principal constitutent direct attachment takes place with a rate of $1 \times 10^{-15} cm^3/sec$.

Processes leading to the destruction of negative ions must include ion-ion recombination, associative detachment, collisional detachment and photo-detachment. The electron affinities of many of the molecules which result in negative ions are unknown at present. The affinity of O is reasonably established at 1.5 ev, Schulz (1962). A lower limit of 0.15 ± 0.05 ev was set by Burch et al (1958) in order to explain photodetachment cross section data of O_2^-. Phelps and Pack (1966) obtained a value of 0.43 ± 0.02 ev. The affinity of CO_4^- is 1.22 ev Pack and Phelps (1966). The affinities of CO_3^- and CO^- are unknown. One would expect that visible light would not result in detachment of electrons from CO_3^- and CO_4^-. In addition to needing shorter wavelengths one must consider the reduced intensity which occurs at Mars.
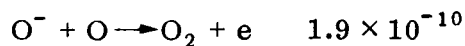
Associative detachment can be an important loss process particularly when O and CO are present. The reaction

$$CO + O^- \longrightarrow CO_2 + e$$
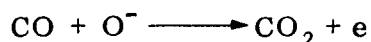
has a rate of $8 \times 10^{-10} cm^3 \ sec^{-1}$, Moruzzi and Phelps (1966). In addition the reactions

$$O_2^- + O \longrightarrow O_3 + e \qquad 3.3 \times 10^{-10}$$

and

$$O^- + O \longrightarrow O_2 + e \qquad 1.9 \times 10^{-10}$$
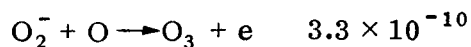
must be considered. For rate coefficients of this order negative ions will be destroyed for all but the smallest concentrations of O.

Collisional detachment was not found to be appreciable in O_2-CO_2 mixtures where CO_2 is the major constitutent. However the reaction
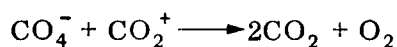
$$CO_4^- + CO_2 \longrightarrow 2CO_2 + O_2 + e$$

does occur at high pressures Mourzi and Phelps (1966).

The rate coefficients for such ion-ion recombination processes as

$$CO_3^- + CO_2^+ \longrightarrow 2CO_2 + O$$
$$\longrightarrow CO + CO_2 + O_2$$

and

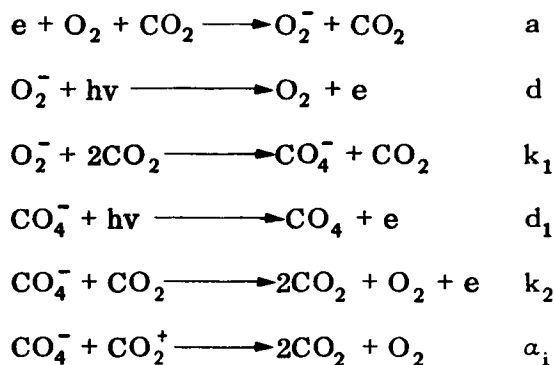$$CO_4^- + CO_2^+ \longrightarrow 2CO_2 + O_2$$
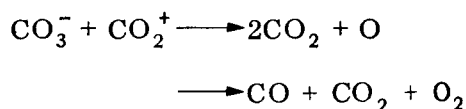
as well as similar negative ions reacting with CO^+ and O_2^+ are unknown at present.

On the basis of the abundance of constituents of the Martian atmosphere, the following negative ion reactions may be selected as being the most important

$$e + O_2 + CO_2 \longrightarrow O_2^- + CO_2 \qquad a$$

$$O_2^- + hv \longrightarrow O_2 + e \qquad d$$

$$O_2^- + 2CO_2 \longrightarrow CO_4^- + CO_2 \qquad k_1$$

$$CO_4^- + hv \longrightarrow CO_4 + e \qquad d_1$$

$$CO_4^- + CO_2 \longrightarrow 2CO_2 + O_2 + e \qquad k_2$$

$$CO_4^- + CO_2^+ \longrightarrow 2CO_2 + O_2 \qquad a_i$$

Ion – ion recombination will become important only during disturbed events and can be neglected in the present analysis. The resulting equilibrium expression for the ratio of CO_4^- ions to electrons, N_e, is

$$\lambda_{CO_4^-} = \frac{a \, k_1 \, [O_2] \, [CO_2]^3}{\{d + k_1 [CO_2]^2\} \, \{d_1 + k_2 [CO_2]\}}$$

Molecular oxygen, while not having been detected spectroscopicly, is probably a minor constituent in the atmosphere. Berkner and Marshall (1967) estimate that the upper limit is 1% of the total atmosphere. The E region model assumes the concentration of O_2 to be 10^{-4} of the total number density. Figure 3 gives $\lambda_{CO_4^-}$ as a function of CO_2 number density for an assumed O_2 concentration of $10^{-2} \, [CO_2]$. A value of $2 \times 10^{-17} cm^3 \, sec^{-1}$ will be taken for k_2 and $0.2 \, sec^{-1}$ will be chosen for the photodetachment rate of O_2^-. Photodetachment from CO_4^- will be neglected.

The dashed curve corresponds to the nighttime case of no photodetachment. In such circumstances

$$\lambda_{CO_4^-} = a \, [O_2]/k_2 = 10^{-15} \, CO_2$$

The altitude for $\lambda = 1$ can occur at about the same altitude as on Earth for the E region model. For an F-2 model $\lambda = 1$ almost 20 km below the corresponding E region altitude. In any event, for these assumptions $\lambda = 1$ for $[CO_2] = 10^{15}$ cm^{-3}.

DISTRIBUTION OF IONS AND ELECTRONS

The results of the previous sections can now be combined to estimate the ion density distribution with altitude employing the sources of ionization and loss processes discussed earlier. In Figure 4 is shown the lower Martian ionosphere as it would appear if the observed electron density distribution is that of an E region. The observed electron density distribution will decrease rapidly to a minimum of the order of 100 cm^{-3} at 85 km. Below this altitude the positive ion density increases due to cosmic ray ionization, the minimum represents the matching of the solar photoionization rate with that of the cosmic rays. Ionization increases to the planet's surface where the ion density is of the order of 3×10^3 cm^{-3}. Also shown is the negative ion, CO_4^-, density corresponding to the curve of λ given in Figure 3. Because of the reaction between CO_2^+ and O_2 the major ion of the cosmic ray layer will be O_2^+ even if O_2 is only 10^{-4} of the total number density.

Figure 4b shows the distribution which would be expected if the ionosphere is an F-2 layer. In this instance the minimum will occur in the vicinity of 50 km. The species of ion will be the same as in the E region case. The effect of meteors has not been included in these ion density estimates. In the case of the E region model the deposition heights are such that the meteors contribute mostly in the height range where the maximum of ionization is observed. In

the F-2 case the deposition will occur in the 75 to 85 km range — see Table II.

PARTICULATE MATTER IN THE ATMOSPHERE OF MARS

The existence of a particulate matter on Mars is well established. Öpik (1966) has reviewed the optical observations of the planet and concludes — "Limb darkening at full phase in green, yellow, and red light indicates absorption by atmospheric haze, aerosols, and dust." An average particle radius of 10^{-3} cm was deduced from the lifetime of the large (area 3×10^3 km) yellow cloud which was observed in 1956. Greenspan (1965) has postulated that concentrations of 10^8 to 10^9 ice particles/cm^2 column and of 0.2μ diameter are needed to bring the visual, polarimetric, and ultraviolet photometric estimates of the surface pressure of Mars into agreement. The most dramatic evidence for the existence of particulate matter in the atmosphere of Mars is the photograph of the limb of the planet taken by the TV equipment aboard Mariner IV, Leighton et al (1965) and Leighton (1966). A cloud estimated to extend to an altitude of 100 km was visible in one of the photographs. Nicks (1967) has suggested that the cloud might be composed of dry ice crystals. A similar suggestion was put forward by Hess (1950) in order to explain the haze layer. The F-2 model of Fjeldbo et al (1966a,b) has mesopheric temperatures which fall in the range where dry ice crystals would be formed.

The attachment of electrons and ions to micron and submicron sized particles has been observed in the laboratory and the terrestrial atmosphere. Formulae for the attachment of charged particles and the resulting charge on

the dust particles have been derived by several authors, Fuchs (1947), Gunn (1954), Bricard (1962, 1965), and Whipple (1965).

One of the elements involved in thunderstorm formation is the charging of cloud particles see e.g. Shishkin (1965). Such charging should also occur within the clouds observed on Mars. The net effect of particulate matter in the Martian atmosphere will be a reduction in the electron and ion density distribution. The existence of such particle densities at altitudes 20 km above the cloud top is a possibility. Attachment to dust should be considered as a potential loss process for ions.

DUSTSTORMS

It is noticed that yellow clouds move frequently in the vicinity of the thermal equator of Mars. Their duration is usually only one to three days and for this reason they are termed "duststorms". Newbauer (1966) places the origin of the clouds in thermal convection which leads to the formation of dust devils. The speed of the wind necessary to initiate the storm has been estimated to be 145 km/hr and may exceed 300 km/hr, 100 km/hr is frequently observed Gifford (1964). Such storms are of interest in the study of atmospheric electricity on the planet. For instance Chalmers (1965) has noted that terrestrial duststorms give rise to large electric potential gradients and has noted a case where lightening was observed.

CONCLUSION AND RECOMMENDED EXPERIMENTS

Recent data on the atmosphere and ionosphere of Mars has been utilized to outline the electrical properties of the lower Martian atmosphere. Suggested

sources of ionization include; solar ultraviolet and x-radiation, cosmic rays, meteors, radioactive material within the atmosphere and radioactivity of the surface, the solar wind, and protons originating from solar flares.

Loss processes including charge rearrangement reactions, dissociative recombination of ions and electrons and negative ion formation have been detailed. The principal negative ion may be CO_4^- if sufficient O_2 is present. The negative ion CO^- is also a possibility.

The altitude distribution of the cosmic ray produced ion density distribution is shown to decrease from a maximum of 3×10^3 cm^{-3} at the surface to less than 10^2 cm^{-3} between 55 and 90 km depending on atmospheric model. The tail of the solar photoionization rate when combined with this cosmic ray production will lead to the formation of an ion density minimum. The magnitude of the density and location of the minimum will depend on the atmospheric model chosen, the activity of the sun and solar zenith angle as well as such additional causes as the presence of particulate matter and meteor ionization. This minimum will disappear for Fregian models if solar Lyman alpha plays any role in the ionization of the Martian atmosphere. Such an effect could occur if nitric oxide is a constitutent of the Martian mesosphere. If no negative ions are present, the electron density distribution will closely resemble the positive ion density curves given in figure 4. The electron density will be less than 5×10^3 cm^{-3} at altitudes below 50 km.

The available data on the Martian ionosphere has lead to several different interpretations of the formation of ionosphere, none of which is without objections. It is clear that additional experiments will have to be performed before

the uncertainty can be resolved. It has been suggested that additional data be obtained from an orbiting vehicle. Performance of the occultation experiment at smaller solar zenith angles would certainly be useful since under conditions the ionization at lower altitudes will be enhanced allowing one to obtain more information at lower altitudes. It is such low altitude data which holds the solution to the problem of what type ionospheric layer was observed by Mariner IV. In order to investigate the lower ionosphere of Mars it is necessary to launch probes into the atmosphere. Such probes should be instrumented to measure concentrations of positive and negative ions as well as their species. An experiment should be carried out to monitor the conductivity of the Martian atmosphere near the surface of the planet on an extended basis.

Recent planetary probes to Venus have determined that this planet's atmosphere is also largely CO_2 so that the results of the present study can be carried over to that atmosphere taking into account, of course the greater pressures and higher temperatures encountered on Venus. Vinogradov et al (1968) have indicated that the Venusian atmosphere contains water vapor. The presence of sufficient quantities of water molecules in the Martian and Venusian atmospheres can change the ion species by forming hydrated molecules such as $CO_4 \cdot (H_2O)_n^-$ and positive ions of the type $H_5O_2^+$ and H_3O^+.

FIGURE CAPTIONS

Figure 1.   Electron concentration vs height profile observed with two frequency
occultation experiment aboard Mariner IV.

Figure 2.   Ion pair production function resulting from solar ultraviolet and
x-radiation and galactic cosmic radiation E and F-1 solar production functions
are from McElroy (1967).

Figure 3.   Negative ion to electron ratios for different atmospheric models.

Figure 4.   The lower ionosphere of Mars for a) E-region model and b) F-2
region model.

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ELECTRON NUMBER DENSITY cm^{-3}

ALTITUDE km.

CORRESPONDING
ALTITUDE
IN km
FOR ATMOSPHERIC
MODELS

NEGATIVE ION/ELECTRON DENSITY (λ)

PARTICLE DENSITY cm⁻³

JOHNSON F-2

McELROY E

McELROY ATMOSPHERE

JOHNSON ATMOSPHERE

ALTITUDE~Km

NUMBER DENSITY~cm^{-3}

Ne

Ne

N+

N-